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THE ROLE OF TURBULENCE IN THE TRANSPORT OF AN ATMOSPHERIC EDDY

[This is a translation of an article written by L.Yu. Ryshakov in Problemy Arktiki i Antarktiki (Problems of the Arctic and Antarctic), No. 1, Leningrad, 1959.]

The idea of conceiving atmospheric movements as eddies of varying dimensions appertains to the eminent Russian meteorologist P.I. Brounov and was formulated by him at the end of last century [3]. Nowadays this concept has found universal recognition, inasmuch as it establishes the likeness of such atmospheric phenomena as planetary eddies, active centers of the atmosphere, moving cyclones and anticyclones, local circulation, etc.

An important problem of modern meteorology is the problem of the genesis, development and motion of cyclones and anticyclones which constitute one of the fundamental links of the general atmospheric circulation. Enormous masses of air take part in the motion within the cyclone-anticyclone system. The life cycle of theses eddies is rather complex. The predetermination of the variation in intensity of cyclones and anticyclones, their trajectories and the velocity of vertical propagation of the disturbances is very difficult because it depends on many factors.

Present-day research on the dynamics of large-scale atmospheric processes is based on the eddy-transfer equation (Fridman's equation). Upon analyzing this equation, L.T. Matveyev [7] formulated rules for a qualita-

ve evaluation of the conditions for the origination, intensification d motion of cyclones and anticyclones. Let us write the eddy-transfer uation in the following form /4, ch.17/:

$$\frac{\partial \Omega_z}{\partial t} = -\left(U\frac{\partial \Omega_z}{\partial x} + V\frac{\partial \Omega_z}{\partial y}\right) + \frac{2\omega_z}{T}\left(U_g\frac{\partial T}{\partial x} + V_g\frac{\partial T}{\partial y}\right) - \beta V_N - $$
$$- (2\omega_z + \Omega_z)\left(\frac{\partial U}{\partial x} + \frac{\partial V}{\partial y}\right) + \frac{1}{\rho}\frac{\partial}{\partial z}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right). \qquad (1)$$

Here, U and V are the components of the wind velocity on the axes x d y, respectively (the x-axis points eastward, and the y-axis north- rd); Ω is the vertical component of the relative vorticity; β is the riation with latitude of Coriolis' parameter: ρ is the air density; id τ_x, τ_y are the components of the shearing stress of the turbulent rces.

It ensues from Eq. (1) that the variation of the vertical component the vorticity in a certain region is determined by the following fac- ors:

A. The horizontal transport (advection) of an eddy from adjacent re- ons (1st addend). This addend is positive in the case of predominantly clonic vorticity of the air stream transported; it is negative if the orticity is predominantly anticyclonic.

B. The geostrophic advection of heat or cold (2nd addend). In the se of the former, horizontal baroclinicity contributes to the formation d subsequent intensification of a negative (anticyclonic) eddy, and

in the case of the latter - of a positive (cyclonic) eddy.

C. Meridional eddy motions (3rd addend). The motion of an eddy having an equatorially directed component produces an effect which tends to intensify a cyclonic eddy and to weaken an anticyclonic one; a polewardly directed meridional component provokes the opposite effect.

D. Horizontal divergence of the wind velocity (4th addend). A positive divergence tends to intensify an existing anticyclonic eddy, whereas a negative divergence (convergence) tends to intensify an existing cyclonic eddy.

E. Turbulent friction (5th addend, which will be discussed in more detail below).

In ref. /8/, L.T. Matveyev and V.A. Zvyabrikov, using factual data relating to 46 cases of eddy formation, calculated the following quantities: 1) the individual variation of the eddy in geostrophic approximation from the Formula

$$\left(\frac{d\Omega_z}{dt}\right)_g = \frac{1}{2\omega_z\rho} \cdot \frac{d\Delta P}{dt},$$

Where ΔP is the pressure Laplacian; 2) the geostrophic temperature advection; 3) a factor taking account of the latitude effect. The calculations were effected at sea level and pressures of 850, 700 and 500 mb. Analysis of the mean values given in the reference work shows that the above quantities diminish with altitude: notably, at 500 mb each of them

decreases on the average by one-half as compared to the ground value. The baroclinic term in Eq. (1) has the same order of magnitude as the individual variation of the eddy with time and the addend accounting for the altitude effect. As regards the divergent addend, the authors /8/ state a number of reasons in support of the view that its effect in the process of eddy formation is inconsiderable. Unfortunately, it is not possible to calculate this addend with a sufficient degree of accuracy. Thus, Ye. P. Borisov /1/ showed that when calculating the wind velocity divergence on barometric data from topographic maps, then, at differently directed axes of the coordinates and various values of the differentiation step, the results obtained may differ twice as much for one and the same point.

By calculating the addends of Eq. (1) the author of the present paper investigated the process of intensification and propagation of an anticyclone which on February 1st 1956 was located over Scandinavia and had a pressure of 1047 mb in the center. During the period from 1-4 February this anticyclone shifted to the issue of the river Ob. The pressure in its center rose to 1058 mb. While the anticyclone was moving eastward, the pressure tendency at individual points of its trajectory reached 22 mb per day. The region comprising Europe and West Siberia north of 47° of latitude as well as the North Sea, Norwegian Sea, Barents Sea and Kara Sea was covered with a network of points, for each of which we calculated the values for the vorticity in a geostrophic approximation and the three addends from Eq. (1) determining the advective transfer of the eddy, the temperature advection and the latitude effect.

The calculations were performed for each day and for all levels down to a pressure of 200 mb. Comparison of the values for the addends calculated at the different levels confirmed the view that, together with the emperature advection, the baroclinic term plays a determinative part in the local variation of the eddy. It appeared that at a level relating to isobaric surfaces of 300 and 200 mb the values of the addends accounting for temperature advection frequently exceed the corresponding values in the middle troposphere. Comparison of the aggregate values for the three addends at various levels with the ground values of the diurnal pressure tendency justified the conclusion that under certain conditions the variation of ground-level pressure is determined basically by the processes of vortical and tem perature advection (allowance made for the latitude effect) in the upper troposphere and in the lower stratosphere.

Such a conclusion signifies that there are two layers in the troposphere in which the processes of eddy formation develop most actively, viz., the lower troposphere and the layer that encloses currents. It is natural to assume that vertical turbulence generated in the active layer is bound to involve the adjacent layers in the motion. Indeed, the transport in the atmosphere of any substance as well as of momentum proceeds, mainly, by turbulent interchange. It is known that in the bottom layer of the atmosphere the shearing stress of the turbulent frictional forces decreases with altitude. The action of theses forces also operates to weaken a cyclonic or anticyclonic eddy. However, in the free atmosphere, the forces provoked by turbulence do not disappear, but rather

assume a new and important role.

P. I. Brounov /2/ held that it was necessary to study not only the translational motion of atmospheric eddies within the system of currents of the general circulation of the atmosphere, but also the conditions of transfer of the vortical disturbances from layer to layer.

Academician N.Ye. Kochin /5, 6/ first examined the effect of turbulent frictional forces on atmospheric movements of wide scope. In designing a model of the general atmospheric circulation, N.Ye. Kochin conceived the whole troposphere as a boundary layer and the individual cyclones and anticyclones as turbulent eddies. It was demonstrated that frictional forces materially affect the nature of air currents not only in the bottom layer, but also in the free atmosphere. N. Ye. Kochin proved that a theory postulating a stationary zonal circulation with meridional and vertical wind-velocity components may only be evolved provided the forces of turbulent friction are allowed for.

Let us transform the 5th addend in Eq. (1)

$$\left(\frac{\partial \Omega_z}{\partial t}\right)_\tau = \frac{1}{\rho}\frac{\partial}{\partial z}\left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}\right), \qquad (2)$$

using the Expressions found for the components of turbulent tangential stress

$$\tau_x = k\rho\frac{\partial U}{\partial z}.$$

and

$$\tau_y = k\rho \frac{\partial V}{\partial z}, \qquad (3)$$

where k is the coefficient of turbulence. Neglecting the variations in density and in the coefficient of turbulence in the horizontal plane, we obtain the following Expressions for the components of the shearing stress of the eddy:

$$\frac{\partial \tau_x}{\partial y} = k\rho \frac{\partial^2 U}{\partial y\, \partial z},$$
$$\frac{\partial \tau_y}{\partial x} = k\rho \frac{\partial^2 V}{\partial x\, \partial z} \qquad (4)$$

Let us substitute Expression (4) into Formula (2):

$$\left(\frac{\partial \Omega_z}{\partial t}\right)_\tau = \frac{1}{\rho}\left[\frac{\partial}{\partial z}\left(k\rho \frac{\partial^2 V}{\partial x\, \partial z}\right) - \frac{\partial}{\partial z}\left(k\rho \frac{\partial^2 U}{\partial y\, \partial z}\right)\right]. \qquad (5)$$

Introducing the vorticity, we get

$$\left(\frac{\partial \Omega_z}{\partial t}\right)_\tau = \frac{1}{\rho} \cdot \frac{\partial}{\partial z}\left(k\rho \frac{\partial \Omega_z}{\partial z}\right). \qquad (6)$$

The turbulent stream of the vorticity in the direction z is determined by the Expression (4, ch. I)

$$\rho \overline{\Omega'_z W'} = -\rho k \frac{\partial \Omega_z}{\partial z}. \tag{7}$$

where $\overline{\Omega'_z W'}$ is the averaged product of the eddy fluctuations and the vertical velocity.

Consequently, according to Eq. (6), the formation of eddies under the influence of frictional forces is determined by the derivative with respect to height from the vertical turbulent stream of the eddy. For a qualitative analysis of the conditions of variation of the eddy, let us differentiate over z in Eq. (6); we then obtain

$$\left(\frac{\partial \Omega_z}{\partial t}\right)_\tau = k \frac{\partial^2 \Omega_z}{\partial z^2} + \left(\frac{\partial k}{\partial z} + \frac{k}{\rho} \cdot \frac{\partial \rho}{\partial z}\right) \frac{\partial \Omega_z}{\partial z}. \tag{8}$$

It ensues that the velocity of transport of the eddy in the vertical direction is determined by the peculiar distribution along the vertical Ω_z. Thus, at the origination of a high-altitude cyclone the perturbation is transmitted downward more rapidly in those layers where the gradient Ω_z is larger in terms of absolute magnitude and where this gradient falls off more abruptly from one level to another. On the other hand, the formation of an anticyclonic eddy at high altitude is conducive in due course to the origination or intensification of anticyclonic circulation in the next lower layer in a spot where (the vertical gradient Ω_z is larger and rises) more quickly in a downward di-

rection. In addition, the velocity of vertical transport of the eddy is affected by the degree of turbulence of the atmosphere. The perturbation is transmitted to the neighboring layers at a speedier rate when the coefficient of turbulence and its vertical gradient are large.

In order to illustrate the effect of turbulent frictional forces on the vertical propagation of an eddy, we plotted contour curves tracing the values of the vorticity and coefficient of turbulence for the case of the ~~genesis~~ of an anticyclone on 3 February 1956 in the region of Novaya Zemlya and of a cyclone on 14-16 November 1957 over the Barents Sea (Fig.1). The vorticity was calculated in geostrophic approximation, while the coefficient of turbulence was found from Matveyev's formula.

Up to recently, there existed no method for calculating the coefficient of turbulence in the free atmosphere. This opportunity was only afforded after the publication of L.T. Matveyev's works /9, 10/, in which the following formula was proposed:

$$k = \frac{c^2}{\beta}\,[2{,}314\,\lg\beta - 1{,}157\,\lg(\gamma_a - \gamma) - 0{,}072], \qquad (9)$$

where

$$\beta = \sqrt{\left(\frac{\partial U}{\partial z}\right)^2 + \left(\frac{\partial V}{\partial z}\right)^2}$$

is the vertical gradient of the wind velocity in m/sec. km;

$\gamma = -\partial T/\partial z$ is the vertical temperature gradient in $^{\circ}C/km$;

γ_a is the dry adiabatic temperature gradient;

c is the mean wind velocity in the layer for which k is being determined.

Formula (9) expresses the turbulence of the free atmosphere as a function of its thermodynamic state. This function was found in the course of processing the factual data on overloads for an airplane in free horizontal flight and from data on temperature and wind soundings. The flights were undertaken with airplanes equipped with accelerographs at altitudes of 1 – 2 and up to 12 km. This provides a basis for using Formula (9) for calculations of the coefficients of turbulence throughout the stratosphere and the lower troposphere.

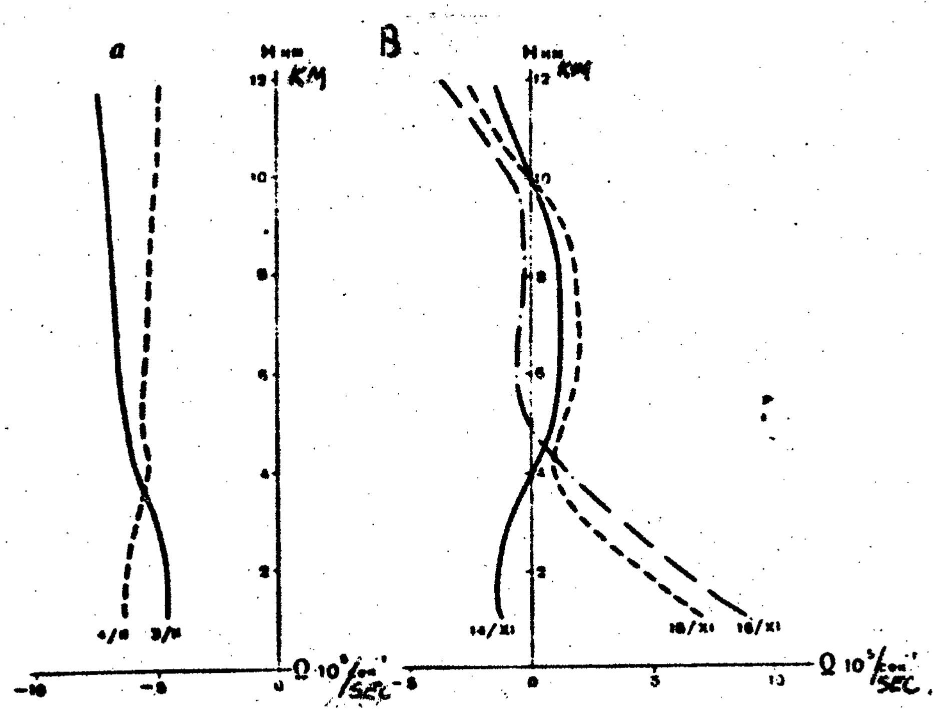

Fig. 1. Development with height of the vertical component of the vorticity in the genesis of a) an anticyclone and b)a cyclone

Table 1 gives the values of the coefficients of turbulence calculated on barometric data from topographic maps for layers situated between levels of 850, 700, 500, 300 and 200 mb. The first of the four cases considered, relates to the period of active anticyclone genesis in February 1956 (tabulated are the values of the coefficient of turbulence over the central region of the anticyclone on 3 February.

Table 1

Coefficients of turbulence on barometric data from topographic maps

1 Слои атмосферы (мб)	3 II 1956 г. 73° 2 с. ш., 40° 3 в. д.				14/XI 1957 г. 75° с. ш., 30° в. д.				15 XI 1957 г. 75° с. ш., 30° в. д.				16/XI 1957 г. 68° с. ш., 50° в. д.			
	γ	β	c	k	γ	β	c	k	γ	β	c	k	γ	β	c	k
850—700	2,0	4,0	8,0	4,4	4,0	7,5	13,0	2,4	6,0	6,3	19,0	61,0	3,0	7,6	20,0	49,2
700—500	6,0	5,0	11,0	20,6	6,0	3,5	21,0	5,7	7,0	6,9	22,0	94,4	8,0	4,2	25,0	150,0
500—300	7,0	2,5	16,0	30,7	6,0	0,9	23,0	2,0	4,0	2,5	28,0	11,0	4,0	2,3	29,0	4,0
300—200	—	—	—	—	0,4	1,8	23,0	2,0	0,0	3,4	30,0	18,0	0,0	3,6	28,0	23,5

1-Atmospheric layers (mb); 2-N. lat.; 3-E. long.

Note: γ is given in °C/km, β in m/sec • km, c in m/sec., k in m^2/sec.

Table 2 presents the values of the factors of eddy formation at various levels on the same day. Whereas conditions in the lower part of the troposphere (data at ground level and at 850 mb) are apt to intensify an anticyclonic eddy, the temperature advection factor at levels of 300 and 200 mb indicates favorable conditions for its abatement. However, on 4 February the intensity of the anticyclonic eddy lessened not only in the upper troposphere and in the lower stratosphere, but also at levels of 500 and 700 mb (Fig.1). Both at 850 mb and at ground level the anticyclonic eddy intensified. The answer to the question as to why the

abatement of the anticyclonic eddy spread to the middle troposphere, but failed to reach the ground layers is to be found in the distribution with height of the coefficient of turbulence on 3 February. In effect, it appears that strong turbulent currents had developed in the atmosphere at levels upward of 700 mb. Under these conditions the influence of frictional forces on the process of eddy transfer proves to be substantial. It is imagined that the weakening of the anticyclonic eddy in the middle troposphere may be explained as a result of the perturbation originated at the higher levels. The layer of 850 - 700 mb, by contrast, may be described as an "arresting" layer, since here the moderate growth of turbulent interchange shielded the lower troposphere from the influence of the higher layers.

Table 2

Factors of eddy formation according to barometric data from topographic maps

Factors of eddy formation	3/II 1956 г. 73° с. ш., 40° в. д.	14/XI 1957 г. 75° с. ш., 30° в. д.	15/XI 1957 г. 75° с. ш., 30° в. д.	16/XI 1957 г. 68° с. ш., 50° в. д.
$\frac{2\omega_z}{T}\left(U_g \frac{\partial T}{\partial x} + V_g \frac{\partial T}{\partial y}\right) \times 10^{11}$ sec^{-2}				
At sea level	1,2	−4,7	−8,6	−3,5
" 850 mb	0,7	−3,8	−6,3	−5,4
" 700 "	1,8	−2,1	−1,1	−0,4
" 500 "	−2,1	0,5	2,2	3,5
" 300 "	−2,6	1,3	1,8	2,4
" 200 "	−1,4	0,7	1,4	0,2
$-\beta V_N \cdot 10^{11}$ сек$^{-2}$				
На уровне моря	0,6	—	6,3	1,6
" 850 mb	0,03	—	3,4	0,4
" 700 "	0,01	—	1,7	0,2
" 500 "	—	—	—	0,2
" 300 "	—	—	—	—
" 200 "	—	—	—	—

Tables 1 and 2 also furnish data relating to the process of cyclone genesis observed over the Barents Sea on 14-16 November 1957. On 14 November observations revealed a weak pressure field at ground level in the region of the Barents Sea. As could be concluded from the barographic maps, this region lay to the left of the axis of an intensive current (with a maximum velocity up to 220 km/hr at 500 mb) proceeding from the northwest in the region of the Greenland Sea toward the region of the middle Volga in the southeast. As seen from the map, on 15 November at 3 a.m. a strong, widely extended cyclone had developed over the Barents Sea. The pressure in the center of the cyclone, at sea level, amounted to 982 mb, as against 1013 mb measured at this point 24 hours previously. An intensive cyclonic formation was observed over the ground center at a level of 850 mb, but later on this perturbation faded rapidly with height. Twenty-four hours later, on 16 November, the ground center of the cyclone had moved 1100 km to the southeast. Pressure in the center of the cyclone fell off further by 10 mb. The magnitude of the positive eddy increased appreciably at a level of 850 mb (Fig.1). The cyclonic reconstruction of the pressure field at levels of 700 and 500 mb proceeded at perceptibly decelerated rates. Above the 500 mb level an intensification of the anti-cyclonic eddy was noted. Thus, despite the origination of a powerful cyclonic eddy in the lower troposphere, its transfer to the higher layers was rather limited. This phenomenon can be explained by way of analysis of the conditions of turbulence in the atmosphere. It appears that during all the days under examination the coefficient of turbulence was

high in the lower and middle troposphere, but was low in the upper troposphere and lower stratosphere.

The examples discussed above show that under certain conditions the frictional forces in the free atmosphere may noticeably affect the transport alike of positive and negative eddying currents in the vertical direction.

The calculated characteristics of eddies and of the coefficient of turbulence make it possible to appraise the order of magnitude of the term $(\partial \Omega_z / \partial t)_\tau$ from Eq. (6). To do this, we first evaluated the vortical flow for various l... from Eq. (7) and thereafter the variation of this flow with height. It was found that the term $(\partial \Omega_z / \partial t)_\tau$ has an order of magnitude equal to 10^{-11} and, hence, is comparable to the values of the baroclinic and latitude terms.

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